Plate Tectonics and the Fission-Powered Engine of the Earth’s Interior:

THE ENDOGENIC HEAT ENGINE AND MANTLE CONVECTION

“The lithosphere is a chemically and mechanically differentiated thermal boundary layer developed in conjunction with thermal convection in the Earth’s mantle.”

Here is perhaps the pithiest and most far-reaching generalization to come out of the plate-tectonic revolution of the 1960s and ’70s. What does it mean?

Why are there so many tectonic plates? Why not just one, as on Mercury or Mars? Has the Earth always had about a dozen major plates? Why do they move at no more than a few centimeters per year with respect to one another? Have they always moved about this fast? What does it mean that decompression melting of mantle rock now yields basalt, whereas it used to yield komatiite, a volcanic rock closer in composition to the magma’s peridotite source rock, during the first half of the Earth’s history?

To begin to answer questions such as these, it helps to think of the Earth as a heat engine, a mechanism that transfers energy from a heat source to a heat sink by way of a working fluid and in the process extracts work (see Lectures 1 and 2). Here we examine the endogenic heat engine of the Earth’s interior, whose principal heat source is radioactive material in the mantle; whose working fluids are the outer core’s molten iron and the sublithospheric mantle’s rock (which behaves as a fluid in response to long-acting stresses); whose heat sink is the exogenic heat engine and ultimately outer space; and whose work is the rearrangement of Earth materials, including the movement and deformation of tectonic plates. We focus on the heat engine’s main component, the mantle.

Heat flow from the Earth’s interior is the power behind the tectonic plates. Where Lecture 4 left off, the Earth was a newly formed ball of red-hot rock. As we saw in Lecture 6, most of the heat now flowing out of the Earth’s interior is accountable to ongoing radioactive decay. Most of it reaches the Earth’s surface through the ocean floor; and it is this part of the heat flow that powers the tectonic plates. This lecture asks how the dwindling energy supply has affected the plate-tectonic machinery’s operation.

Heat Transport in the Earth

Basic Mechanisms — Heat is energy in the form of random molecular motion. It flows by one means or another down a gradient of decreasing temperature. Observed changes in temperature with depth in the atmosphere, ocean, and solid earth are evidence that heat transfer is occurring.

There are three basic mechanisms of heat transport, and each predominates within a different part of the Earth:

1. Advection, the transfer of heat by transfer of the material that contains it. This is the predominant process in fluids: the outer core’s liquid iron, the sublithospheric mantle’s fluid-like rock, the oceans’ water, and the lower atmosphere’s gas. The rate of heat transport is directly proportional to the rate of the material’s flow and, if the material keeps flowing steadily, to the elapsed time.

2. Conduction, the diffusion of heat in atom-to-atom collisions. This is the predominant process within solids: the lithosphere’s rock and inner core’s iron. Thermal conduction is so amazingly slow that, at even cm/year rates, advective flow within the Earth’s interior is enormously more effective in bringing heat to the Earth’s surface.

3. Radiation, the transfer of energy by atoms’ absorption and reemission of thermal radiation. This is the predominant process in the atmosphere’s energy balance, but it is unimportant in heat transfer within the Earth’s interior. There, as in the atmosphere, radiant energy is continually radiated and absorbed; however, for practical purposes, it is trapped indefinitely.
**Thermal Convection** — A “symbiotic association” between conduction and advection, thermal convection is overwhelmingly the most important process in moving the endogenic heat engine’s two working fluids and in conveying heat to the Earth’s surface. Mantle convection is what makes plate tectonics go. In the core, convection is involved in the generation of the Earth’s magnetic field. Convection is also important on much smaller scales in the lower atmosphere, where it makes clouds and thunderstorms, in hydrothermal systems that form ore bodies at midocean ridges and other volcanic centers, and in many organisms’ cooling mechanisms, including our own (since the regulation of convective cooling is basically what hair and clothing are all about).

Convection may be most familiar from water or coffee or soup simmering on a stove. There, the fluid is heated from beneath and cooled from above, both hot and cold thermal boundary layers and plumes form, and the initially chaotic flow organizes itself into convection cells. Convection within a cooling cup of coffee, another familiar example, is more similar to what takes place within the core and mantle. Convection within water or coffee or soup warming in a microwave oven is a still better analogy. There, the fluid is heated from within, as the core and mantle are, and only cold thermal boundary layers and plumes form.

Thermal convection is a consequence of the thermal boundary layers’ gravitational instability, which in turn is a consequence of the fluid’s expansion on heating and contraction on cooling. The buoyancy forces on the boundary layers are what makes thermal convection go. Cold thermal boundary layers, of which the oceanic lithosphere is one, are gravitationally unstable because they are more dense than the warmer fluid beneath; and hot thermal boundary layers are likewise unstable because they are less dense than cooler fluid above.

**Mantle Convection**

**Qualitative Overview** — Loss of heat to the Earth’s surface causes rock in the cold thermal boundary layer (i.e. the lithosphere) to thermally contract, increase in density, and become gravitationally unstable with respect to hotter, less dense, underlying rock. **Radiogenic heating** perpetuates the gravitational instability, causing deeply buried mantle rock to thermally expand, decrease in density, and rise relative to cooler rock above.

Rock in the sublithospheric mantle is viscoelastic, like glass or silly-putty: it transmits seismic waves as an elastic solid, but responds as a viscous fluid to long-acting stresses. The rock behaves as a fluid because its mineral grains slide past one another. In the asthenosphere, a small amount of melt lubricates the grains; in the underlying mesosphere, at still higher temperatures and pressures, atoms diffuse along grain boundaries, giving a similar lubricating effect. In both instances, the rock’s resistance to flow — its **viscosity** — decreases with increasing temperature because the atoms move more readily along grain boundaries, enabling the rock to deform more readily in response to stress.

In the mantle or any similar body, thermal convection will result if large enough gravitational instabilities develop within the fluid. The body’s dimensions, the properties of the fluid (e.g. density, viscosity, thermal expansion), and the rate of internal heat generation determine whether convection will commence. Given the mantle’s size and properties, some form of thermal convection should be inevitable. A variety of evidence suggests that convection is proceeding on several scales simultaneous. The dominant mode appears to be some form of **whole-mantle convection**, in which convection cells extend from the surface down to the core-mantle boundary. However, at least one form of **layered convection** appears to be taking place; for superimposed on whole-mantle convection cells is some form of circulation associated with the asthenosphere and individual tectonic plates.
The mantle has a built-in thermostat: the rate of heat transport is controlled by the rock’s viscosity, which is temperature dependent and nearly uniform with depth. The more radiogenic heat accumulates, the hotter the mantle will become, the more readily the rock will flow in response to thermally generated differences in density, and thus the more rapidly the mantle will convect away the accumulated heat. Conversely, the cooler the mantle becomes, the more slowly the rock will flow, and thus the more time there will be for radiogenic heat to build up, warm the mantle, and decrease the rock’s resistance to convective flow that will carry the accumulated heat to the surface.

The temperature dependence of the mantle’s viscosity leads us to expect that heat has been convected away very nearly as fast as it has been generated. The early mantle should have convected most of the heat generated during the Earth’s formation and differentiation, and should have approached its present-day overall temperature within the first few hundred million years.

A Model Convection Cell — To see how rates of plate movement relate to convective cooling of the Earth’s interior, and to see what factors determine these rates, let us examine the idealized, two-dimensional, whole-mantle convection cell diagrammed in the accompanying figure.

Simplifying Assumptions — The following are appropriate for a layer of Newtonian fluid which is uniformly heated from within, which is insulated on its lower boundary and chilled to a constant temperature (“Surface Temperature”) on its upper one, and which is freely convecting in a steady state:

✓ The convection cell is as wide as the convecting layer is deep (“Cell’s Width” = “Cell’s Depth”), as convection cells tend to be in a freely convecting layer such as described.

✓ Flow lines run parallel to the convection cell’s sides; and instead of running smoothly around the corners, they make abrupt right-angle bends. The fluid completes a circuit of the cell in a certain period termed here the “Cycle Time.” The speed of flow increases linearly from zero at the cell’s center to a maximum at its edges, so that the strain rate within the fluid (the change in flow rate with change in depth) is inversely proportional to the Cycle Time:

\[ \text{Strain Rate} = \frac{4 \cdot \text{Cell’s Depth}}{\frac{1}{2} \text{Cell’s Depth}} \cdot \frac{\text{Cycle Time}}{0} = \frac{1}{\text{Cycle Time}} \]  

(9.1)

Since the shear stress resisting the flow equals the viscosity times the strain rate in a Newtonian fluid equals the viscosity times the strain rate, the shear stress resisting convective flow is thus proportional to the viscosity divided by the Cycle Time:

\[ \text{Shear Stress} \propto \frac{\text{Viscosity}}{\text{Cycle Time}} \]  

(9.2)

✓ The temperature is uniform within the fluid (“Central Temperature”) except within the cold thermal boundary layer (read: oceanic lithosphere) and associated cold thermal plume (read: subducting slab). The temperature gradient is linear within the boundary layer and equal to the temperature difference across it (Central Temperature - Surface Temperature) divided by the boundary layer’s thickness at the particular point.

✓ Heat flow within the boundary layer is exclusively (as opposed to predominantly) in the upward direction. Thus, by Fourier’s Law of Heat Conduction, the heat flow at any point on the cell’s upper surface will be directly proportional to the temperature gradient at that point:
Heat Flow
\[ \Delta \propto \frac{\text{Temperature Difference}}{\text{Layer's Thickness}} \]  

\hspace{1cm} (9.3)

**Heat Flow and the Thermal Boundary Layer** — The cumulative heat loss through any point on the cell’s upper surface will be directly proportional to the boundary layer’s thickness at that point times difference between the temperature of the cell’s interior and the average temperature within the boundary layer:

\[ \text{Cumulative Heat Loss} \propto \text{Temperature Difference} \cdot \text{Layer’s Thickness} \]  

\hspace{1cm} (9.4)

Since the heat flow in (9.3) is simply the time rate of change in the cumulative heat loss as given in (9.4), we have

\[ \frac{\text{Temperature Difference}}{\text{Layer’s Thickness}} \propto \text{Temperature Difference} \cdot \text{Layer’s Rate of Thickening} \]  

\hspace{1cm} (9.5)

\[ \text{Boundary Layer’s Thickness} \propto (\text{Time in Contact With Boundary})^{1/2} \]  

\hspace{1cm} (9.6)

As it applies to oceanic lithosphere, (9.6) becomes the important generalization

\[ \text{Oceanic Lithosphere’s Thickness} \propto (\text{Lithosphere’s Age})^{1/2} \]  

\hspace{1cm} (9.7)

The heat flux out of the convection cell is proportional to the total flow through a point, as given by (9.4) and (9.6), divided by the time it takes the point to traverse the cell’s upper boundary (i.e. (Cycle Time)/4):

\[ \text{Heat Flux} \propto \frac{\text{Temperature Difference}}{(\text{Cycle Time})^{1/2}} \]  

\hspace{1cm} (9.8)

**The Force Driving Thermal Convection** — This force arises from **thermal contraction** of fluid in the cold thermal boundary layer, and it is simply the weight of the cold thermal plume within the surrounding fluid. The density difference between the plume and the rest of the fluid is directly proportional to the temperature difference across the thermal boundary layer:

\[ \text{Density Difference} = (\text{Plume’s Mean Density} - \text{Central Density}) \]  

\[ \propto (\text{Central Temperature} - \text{Plume’s Mean Temperature}) \propto \text{Temperature Difference} \]  

\hspace{1cm} (9.9)

The force driving convection, in turn, is proportional to this density difference times the plume’s length (which does not figure in the proportionality because it is assumed constant) times the plume’s thickness (which is simply the boundary layer’s thickness at the time it begins to subduct, as found from (9.6)):

\[ \text{Driving Force} \propto \text{Temperature Difference} \cdot (\text{Cycle Time})^{1/2} \]  

\hspace{1cm} (9.10)

**The Balance of Forces** — For convection in a steady state, the driving force in (9.10) equals the viscous force resisting fluid flow, which is the shear stress in (9.2) times the convection cell’s perimeter (which does not figure in the proportionality because it is assumed constant):

\[ \text{Driving Force} = \text{Viscous Resisting Force} \]  

\[ \text{Temperature Difference} \cdot (\text{Cycle Time})^{1/2} \propto \frac{\text{Viscosity}}{\text{Cycle Time}} \]  

\hspace{1cm} (9.11a)

\hspace{1cm} (9.11b)
The balance of forces thus establishes one of two fundamental relationships among the variables controlling thermal convection:

\[
\text{Cycle Time} \propto \left[ \frac{\text{Viscosity}}{\text{Temperature Difference}} \right]^{2/3}
\]  

(9.12)

**The Heat Balance** — For convection in a steady state, the heat production within the cell equals the heat flux across the cell’s upper boundary, which establishes the other fundamental relationship among the variables:

\[
\text{Cycle Time} \propto \left[ \frac{\text{Temperature Difference}}{\text{Heat Production}} \right]^2
\]

(9.13)

**Synthesis** — Eliminating the Cycle Time (our stand-in for the rate of sea-floor spreading) from (9.12) and (9.13), we obtain the basic relationship among the counterparts to the temperature difference across the lithosphere, the mantle’s viscosity, and its radiogenic heat production:

\[
\frac{\text{Temperature Difference}}{(\text{Viscosity})^{1/4}} \propto (\text{Heat Production})^{3/4}
\]

(9.14)

**The Present as the Key to the Past** — Let us introduce a final assumption: that the proportionalities above have held even if the mantle’s heat budget has not been balanced. We can then recast (9.14) in the following form for use as a key to the past:

\[
\left[ \frac{\text{Past Temperature Difference}}{(\text{Past Viscosity})^{1/4}} \right] = \left[ \frac{\text{Past Heat Production}}{(\text{Past Heat Production})^{3/4}} \right] = \left[ \frac{\text{Present Temperature Difference}}{(\text{Present Viscosity})^{1/4}} \right] = \left[ \frac{\text{Present Heat Production}}{(\text{Present Heat Production})^{3/4}} \right]
\]

(9.15)

As discussed in Lecture 6, heat production in the mantle has been dominated by four radioisotopes for the last 4 Ga, and is given to a good approximation by

\[
\frac{\text{Past Heat Production}}{(\text{Present Heat Production})} \approx \exp \left( \frac{\text{Age}}{3.6 \text{ Ga}} \right)
\]

(9.16)

**Mantle Rock’s Temperature-Dependent Viscosity** — The viscosity of the sublithospheric mantle is a function of temperature, which means that (9.14) and (9.15) can be further simplified. As in many materials, the viscosity has an inverse exponential dependence on temperature:

\[
\text{Viscosity} \propto \exp \left( \frac{\text{Rheological Constant} \cdot \text{Melting Point}}{\text{Temperature}} \right)
\]

(9.17)
For mantle rock, the dimensionless Rheological Constant is about 30. As determined from studies of ongoing postglacial rebound in the Baltic Sea and Hudson Bay regions, the mantle’s viscosity is a more or less uniform $1 \times 10^{21}$ pascal-seconds except in the asthenosphere, where a tiny fraction of melt (i.e., molten rock) reduces it to as little as $4 \times 10^{19}$ Pa-s. The apparent explanation is that the rock’s temperature closely approaches the melting point of at least one of its minerals throughout the mantle, and slightly exceeds that point in the asthenosphere. In other words, the rock’s homologous temperature (the ratio of its temperature to its melting point, as appears in (9.17)) is very near unity throughout the mantle. Substituting the appropriate values into (9.17) gives this simpler expression for the mantle’s viscosity:

$$\frac{\text{Past Viscosity}}{\text{Present Viscosity}} \approx \exp \left[ 30 \cdot \left( 1 - \frac{\text{Past Central Temperature}}{\text{Present Central Temperature}} \right) \right] \quad (9.18)$$

**Applying the Model** — Taking the mantle’s present mean temperature to be about 2300 K (as estimated from seismological data and laboratory simulations of mantle rock), taking the past and present surface temperature to be about 300 K (as consistent with evidence of abundant water and life throughout geologically recorded history), and making the appropriate substitutions into (9.15), one finds

$$\frac{1 + \frac{\text{Difference in Mantle Temperature}}{2000 \text{ K}}}{\left\{ \exp \left[ 30 \cdot \left( -\frac{\text{Difference in Mantle Temperature}}{2300 \text{ K}} \right) \right] \right\}^{1/4}} \approx \exp \left( \frac{\text{Age}}{3.6 \text{ Ga}} \right)^{3/4} \quad (9.19)$$

**The Mantle’s Mean Temperature as a Function of Time** — Taking logarithms in (9.19), messing with the algebra, and applying the approximation $\ln (1 + x) \approx x$ for $|x| << 1$ where appropriate, we find

$$\text{Difference in Mantle Temperature} \approx (56 \text{ K/Ga}) \cdot \text{Age} \quad (9.20)$$

Despite the many approximations and simplifying assumptions, the predicted rate of cooling is in good agreement with rigorous estimates in the 30-100 K/Ga range. We conclude that the mantle should have been 200-300 K hotter overall when the oldest surviving rocks formed.

The predicted decrease in the mantle’s temperature is borne out by the occurrence of komatiite, the characteristically Archean volcanic rock similar to basalt that we saw in Lab 1, formed through more complete melting of the same mantle source rock at a temperature 200-300 K higher.

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2 “Elementary, my dear Watson.” Here we have what, by the standards of conventional Earth history courses, is the rarest of rarities: a result of fundamental importance obtained by rigorously testing a prediction rigorously derived from the fundamental physics — a geopedagogically all but unprecedented triumph of “Dr. Moriarty-style” applied science brought down to the level of liberal education. But of course this is Cornell, where students are used to such things and faculty in their modesty do not waste class time pointing them out.
Related Trends in Plate Tectonics — Successfully predicting the trend in the mantle’s temperature opens the way to examining related trends in the mantle’s viscosity, in the force driving the tectonic plates, and in the rates of sea-floor spreading and continental drift, as accompanying diagrams show. Plates are predicted to have moved more rapidly in the past. Terranes’ apparent polar wander paths bear out this prediction qualitatively but not quantitatively. The next lecture explains the discrepancy as it turns from the single convection cells considered here to show how the worldwide array of tectonic plates tailors itself to the mantle’s heat-disposal demands.
Study Questions

0. Continuing Questions: How can the deductive, Sherlock Holmes-style approach be used together with the predictive, Dr. Moriarty-style approach to study the operation and evolution of the Earth’s plate-tectonic machinery? (Lectures 1, 2) Can one really get reliable, verifiable information about global-scale processes in the past from a few hand specimens? (Lab 1)

1. Consider the Earth’s mantle as a heat engine. What is its heat source? … its heat sink? … its working fluid? How has its power supply changed through time? Give an example or two of its work.

2. What is heat? What are the basic mechanisms of heat transport? Which are most important in the Earth’s atmosphere? … its hydrosphere? … its lithosphere? … its deep interior?

3. What does it mean to say that the lithosphere is a thermal boundary layer developed in conjunction with convective cooling of the Earth’s interior? How does mantle convection work? What is the driving force and how is it generated? What resists and balances the driving force? What is the power source? Where does the heat end up? What factors determine how fast the sea floor spreads and how fast continents drift?

4. What does it mean to say that the mantle has a built-in thermostat? How does the thermostat work, and how has it influenced the history of the Earth’s cooling and the long-term evolution of plate-tectonic processes?

5. How has mantle convection changed through time? How has the power supply changed? The character of the convecting rock? Rates of sea-floor spreading? How have plate-tectonic processes been affected? … the character of volcanic rocks? … of oceanic crust? … of continental crust? [Hint: refer back to Lab 1]

6. How do meteorites figure in understanding the evolution of plate-tectonic processes? … igneous rocks associated with midocean ridges and subduction zones? … evidence of water and life in the distant past?

7. Is the operation of plate-tectonic processes affected by conditions in the surface environment? If so, how? … by living organisms? If so, how?

8. General as it is, the present treatment of mantle convection applies also to the Earth’s sister planets Venus and Mars during at least some phases in their history. Venus is now experiencing a runaway greenhouse effect in which surface temperatures have risen to as much as 800 K. It also shows surface features resembling the continents, oceans, island arcs, and other features we associate with the lithosphere’s involvement in convective cooling of the Earth’s interior. However, more or less uniform cratering of Venus’ “ocean basins” and other terranes suggests that its version of plate tectonics ceased to operate around 500 million years ago. Could coevolution of Venus’ exogenic and endogenic heat engines account for the extinction of Venusian plate tectonics? If so, how? [These were among the first questions to be asked as the first map of Venus was assembled from the Magellan mission’s radar imagery, and they have yet to be answered to everyone’s satisfaction.]

9. More recently, “magnetic stripes” reminiscent of those on the Earth’s ocean floor have been mapped on what once may have been the floor of Mars’ ocean. There is clear evidence of river channels and now even the shoreline of an ocean left over from Mars’ first billion years or two. Could coevolution of Mars’ exogenic and endogenic heat engines account the extinction of Martian plate tectonics? If so, how? [These are current questions in cutting-edge research.]